FEEDBACK PROCESSES AND CLIMATE RESPONSE

Albert Arking

NASA Goddard Space Flight Center
Greenbelt, MD 20771

ABSTRACT

The repsonse of the climate system to an external perturbation--e.g., a change in solar irradiance or a change in atmospheric opacity due to an increase in CO2--depends rather strongly on feedback processes in the system, which either amplify or dampen the effects of the initial perturbation. A simple representation of the climate system is used to compare several important feedbacks, based upon GCM simulations by various investigators. The models are in general agreement with respect to water vapor feedback, but in wide disagreement with respect to cloud feedback. Because of the arguments raised by Lindzen (1990)--that the processes which determine water vapor mixing ratio in the upper atmosphere are quite different from those which operate in the planetary boundary layer, and that upper tropospheric water vapor might actually decrease even when the boundary layer is getting warmer and more moist--we undertook a study to determine the sensitivity of climate to changes in water vapor at various levels in the troposphere. The result is that climate is just as sensitive to percentage changes in upper tropospheric water vapor, where the mixing ratio is very small, as it is to percentage changes in the boundary layer, which contains the bulk of total column water vapor.

1. INTRODUCTION

This presentation is intended as a summary of what we know and what we do not know about climate feedback and the response of the climate system to external perturbations. The results come mostly from other investigators, the main contribution here being to interpret them within a common framework.

2. CLIMATE FEEDBACK

The importance of climate feedback is well illustrated by looking at successive stages as we allow more and more of the components of the climate system to respond to an external perturbation. For example, if we fix all parameters of the climate system except surface temperature then the change in mean global surface temperature, in response to doubling the CO2 concentration, will be approximately 1.2C. It is a purely radiative process, and there is not much disagreement in this result.

In the next step we allow the temperature profile to change along with the surface temperature. In that case the effect is no longer purely radiative, and the result now depends upon the dynamical processes in the model (e.g., non-radiative energy transport). Allowing the temperature profile to change reduces the change in mean global surface temperature from 1.2C to ~1.0C. This is the *lapse rate* feedback, which is always negative.

TABLE 1. The change in surface temperature (ΔTSFC) due to a doubling of atmospheric CO_2, as each parameter is incrementally allowed to vary.

PARAMETER	ΔT_{SFC} (C)
T_{SFC}	1.2
T(z)	~1
WATER VAPOR	~2
SURFACE ALBEDO	2.5 - 3.0
CLOUDS	2 - 5

We continue in this fashion, with the results shown in Table 1. Allowing the water vapor mixing ratio to change brings in a whole slew of hydrological processes--e.g., surface evaporation, moist convection, dynamic transport, cloud formation, and precipitation. In most models the net result of all of these prcocesses is for relative humidity to remain roughly the same, and the change in mean global surface temperature increases to approximately 2C. This is a fairly substantial positive feedback, which will be discussed further in Section 5. Until the issue was raised by Lindzen (1990) there was not much disagreement on water vapor feedback being large and positive. This is a characteristic of almost all GCMs used to determine equilibrium response to perturbations, and is consistent with Cess *et al* (1989) finding a large measure of agreement amongst many GCMs in the relationship between forcing and response in cloud-free regions.

Where the models do show substantial disagreement is in surface albedo feedback and cloud feedback. Surface albedo feedback, due primarily to changes in ice and snow cover, is positive in all models (warmer temperature, less ice/snow cover, more absorbed solar radiation). But models differ on the magnitude of the effect. There is even wider disagreement on cloud feedback, discussed in Section 4, where even the sign is uncertain. With all the feedbacks in Table 1, the change in mean global surface temperature due to CO_2 doubling, based on GCM simulations, is in the range 2-5C.

3. A SIMPLE REPRESENTATION OF FEEDBACK

The combined effect of several feedback processes is definitely non-linear, even if we make the simplifying assumptions that the model responds linearly to any forcing (external forcing or the internally generated forcing associated with a feedback process) and that the processes are independent of each other. These assumptions lead to the schematic representation of the climate system shown in Fig. 1.

The *gains* in Fig. 1 add linearly as shown, but the *effect* of the gains on climate system response is non-linear, being proportional to the factor $1 / 1 - \Sigma g_i$. Thus, when the sum of the gains is large--i.e., when $\Sigma g_i \rightarrow 1$--each feedback is substantially amplified by the effects of all the other feedbacks. For example, a feedback with a gain of 0.1 has only a 10% effect if the net gain of all other feedbacks is small (i.e., close to zero), but a 33% effect if that net gain is 0.6.

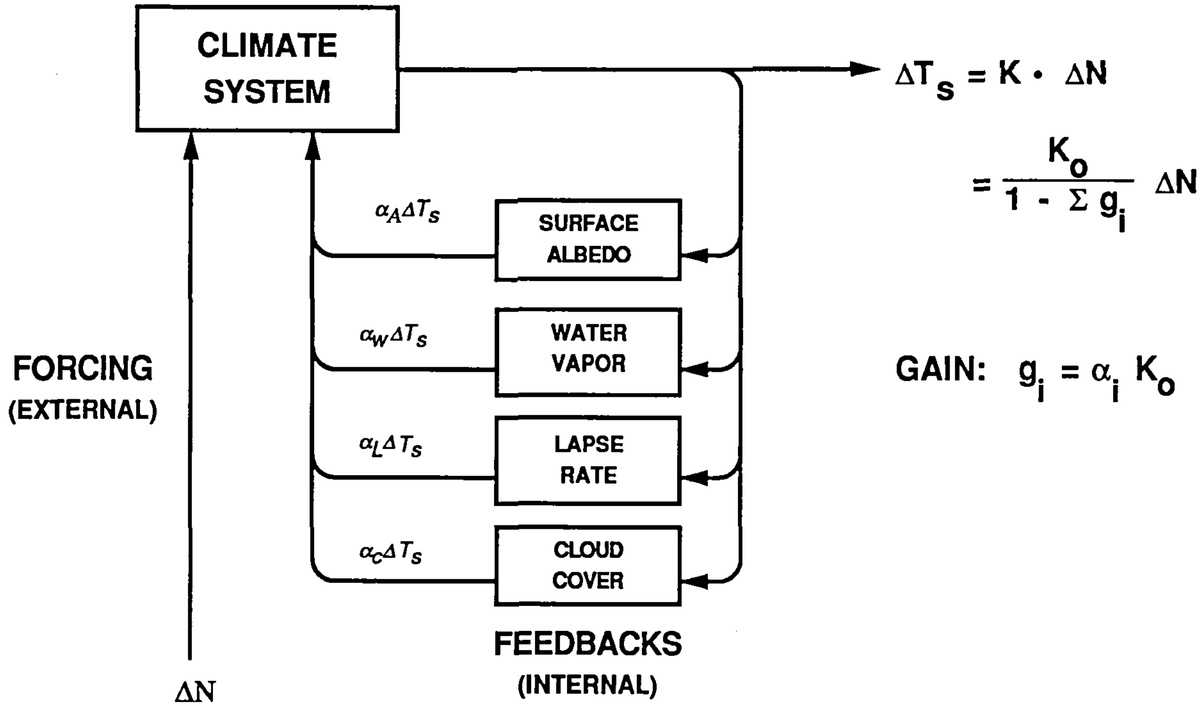

Figure 1. A schematic of the climate system showing how feedback processes contribute to the climate sensitivity coefficient *K*, defined as the change in surface temperature ΔT_s per change in net energy flux at the top of the atmosphere ΔN. The gain *g* associated with any feedback process is non-dimensional; it is the product of α (the change in *N* due to that process, per change in surface temperature) times K_o (the change in surface temperature that would result from an externally induced change in *N* without feedback).

TABLE 2. The calculated gain associated with feedback processes in the GFDL and GISS models.

FEEDBACK PROCESS	GFDL	GISS
Surface Albedo	.16	.09
Water Vapor & Lapse Rate*	.43	.40
Cloud Cover	.11	.22
Total Gain	.70	.71

*Effect of stratospheric response to CO_2 change is subtracted from GFDL results to make it consistent with GISS.

4. CLOUD FEEDBACK

The feedback gains of two of the models used to investigate the effects of doubling CO_2, GISS (Hansen et al, 1984) and GFDL (Wetherald and Manabe, 1988), are shown in Table 2. The mean global surface temperature response is about the same--4.2C for GISS and 4.0C for GFDL--so that they have nearly the same total gain, ~ 0.7. The breakdown of the gains, however, exhibits a large difference in cloud feedback, which is compensated by the difference in surface albedo feedback.

Our simple representation of feedback allows us to play a game with the numbers in Table 2, in which we exchange the cloud parameterization schemes between the two models. The result is an increase in the net gain for the GFDL model to 0.81 and a decrease for the GISS model to 0.60. In that case the response to a doubling of CO_2 in the GFDL model would be 6.3C, and in the GISS model 3.0C. Although this representation of climate system response is highly simplified, it does convey the essential result that cloud processes can have a very large effect on model response.

A further illustration of model sensitivity to the parameterization of cloud processes is provided by the results of a study at the British Meteorological Office (Mitchell *et al*, 1989). By changing the parameterization of the hydrological cycle, including cloud formation processes, and the parameterization of cloud optical properties, the response to doubled CO_2 changed by almost a factor of three (1.9C versus 5.2C). The 1.9C result implies negative cloud feedback.

Cloud feedback now stands as the major cause of uncertainty in climate sensitivity studies. One has to conclude at this point that clouds *may* have a strong influence on climate change, but we are far from knowing the magnitude, and perhaps even the sign, of this influence.

5. WATER VAPOR FEEDBACK

While models used up to now for simulating radiative perturbations, such as the effects of changing the solar irradiance or changing CO_2, are generally in agreement that the water vapor feedback is large and positive, Lindzen (1990) has recently questioned this result. He accepts that water vapor in the boundary layer will increase with increasing temperature, consistent with the models, but argues that water vapor above the boundary layer will *actually decrease.*

It is generally believed that water vapor in the upper troposphere is controlled primarily by moist convection and precipitation, processes which are only crudely parameterized in climate models. It is conceivable, therefore, that changes in upper tropospheric moisture may not be accurately computed in the models. To determine the relative importance of upper tropospheric moisture versus boundary layer moisture, we employed a one-dimensional radiative equilibrium model with convective adjustment. (See Ramanathan and Coakley, 1978, for a review of these models.)

The model calculates an equilibrium temperature profile such that the net radiative flux is zero above a convective zone. Within the convective zone the temperature follows a moist adiabatic lapse rate. The height of the convective zone is adjusted to be the minimum height for which the temperature

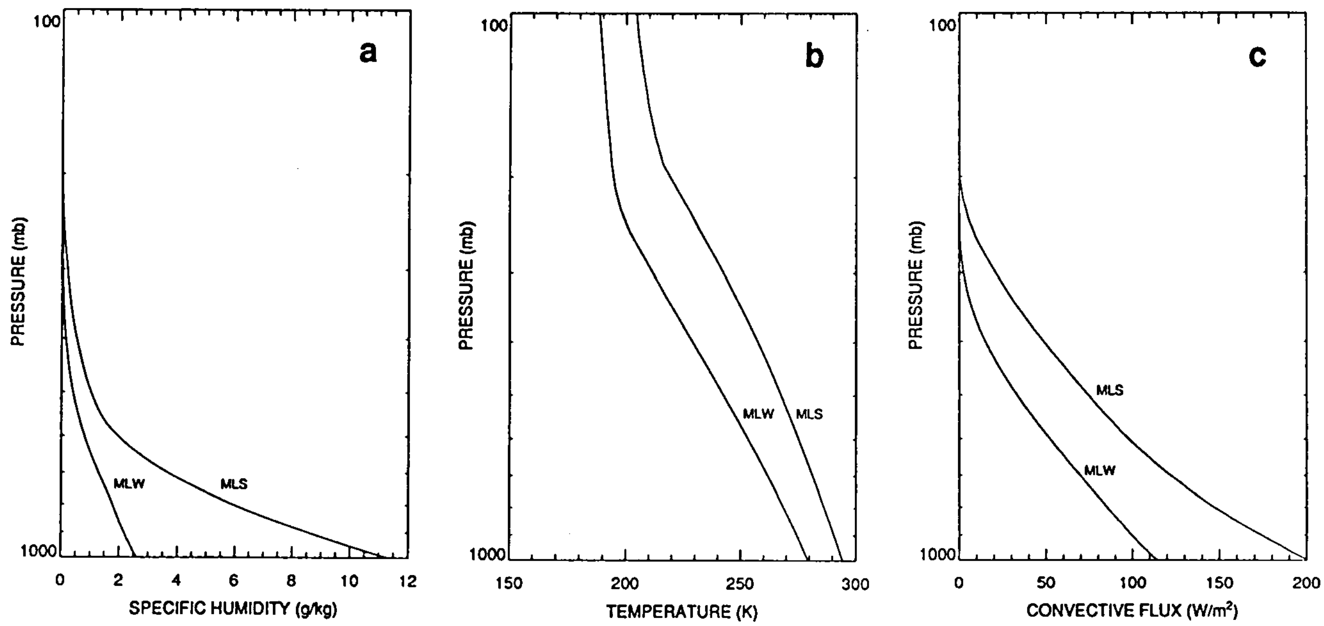

Figure 2. For two humidity models, MLW (mid-latitude winter) and MLS (mid-latitude summer), shown in **a**, the equilibrium temperature (**b**) and the convective flux (**c**). The prescribed parameters are: no clouds, CO2 concentration 330ppm, and net absorbed solar flux 236W/m^2 for MLW and 290W/m^2 for MLS. The height of the convective zone is determined to be 240mb and 180mb for MLW and MLS, respectively.

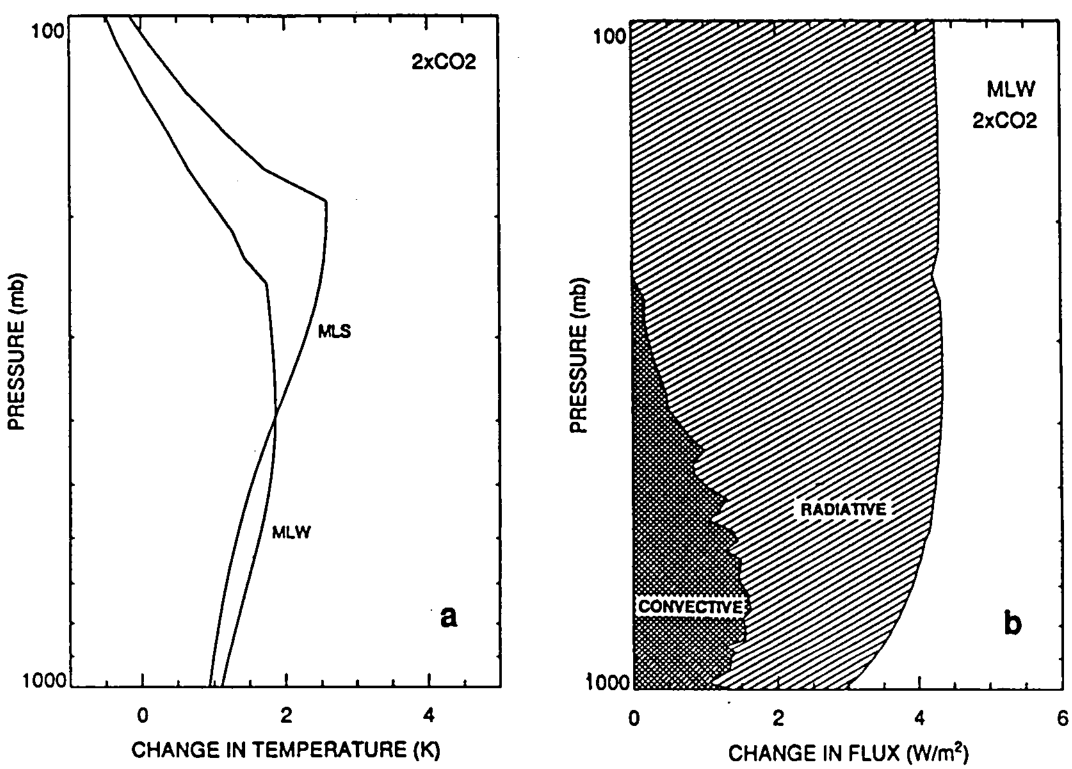

Figure 3. The change in temperature (**a**) and the change in convective and total flux (**b**) required to achieve equilibrium after increasing CO2 from 330ppm to 660ppm. (Note that the irregular shape of the convective flux is an artifact of the numerical procedure and has no physical significance.)

lapse rate above the zone is less than adiabatic. The model has 57 layers from the surface to 10mb, and the radiation routines are based on Chou and Arking (1980 and 1981) and Chou and Peng (1983).

An equilibrium solution obtained with this model is shown in Fig. 2 for two humidity profiles: MLW (mid-latitude winter) and MLS (mid-latitude summer). In this example we omit clouds and specify a CO_2 concentration of 330 ppm and a net absorbed solar flux of $236W/m^2$ for MLW and $290W/m^2$ for MLS. The net upward flux (radiative plus convective) must match the net downward solar flux at every level. Since the model computes the radiative flux, the convective flux is obtained by subtraction (Fig. 2c). The height of the top of the convective zone is found to be 240mb for MLW and 180mb for MLS.

If we introduce a perturbation--e.g., doubling the CO_2 concentration--with neither temperature nor any other parameter in the model allowed to change, there would be a net imbalance in the outgoing flux at the top of the atmosphere. For the case of doubling CO_2, this deficit is shown in Fig. 3b by plotting the change in radiative flux as a function of height due to the perturbation (the outer boundary of the lightly shaded area). Allowing the model to reach a new equilibrium, the change in temperature is shown in Fig. 3a and the change in convective flux in Fig. 3b (boundary of the darkly shaded area). We interpret Fig. 3b as showing the change in total flux and the change in convective flux required to restore equilibrium, with the shaded areas depicting the partitioning of the change in total flux between convection and radiation.

In response to the change in temperature and the change in the magnitude of the convective flux required to maintain a stable lapse rate, one would expect the total amount and distribution of water vapor to change. However, water vapor is a prescribed parameter, since the processes of evaporation, condensation, and transport of water vapor are not included in this model. We therefore introduced changes in water vapor mixing ratio in individual layers and let the model calculate the change in surface temperature. The result is shown in Fig. 4, where we plot the change in equilibrium surface temperature due to a 50% increase of water vapor in a 40mb layer as a function of the height of the layer in which water vapor is perturbed. For both the MLS and the MLW water vapor profiles the

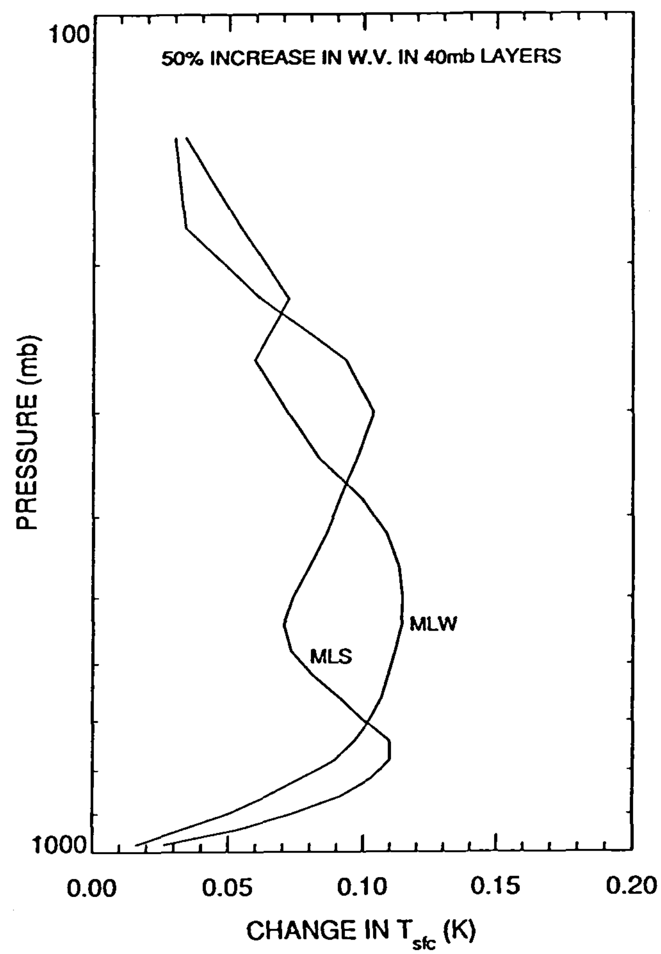

Figure 4. The change in surface temperature due to a 50% increase in water vapor in a 40mb layer, as a function of the height of the layer.

sensitivity varies by less than a factor of two over the range 200-950mb. (The sensitivity must necessarily be zero at the surface.)

From these results we draw the conclusion that the response of the climate system to a percentage change of water vapor is almost independent of the height at which the change takes place. This is so, even though the water vapor mixing ratio varies by a factor of 500 to 1000 within the troposphere. These results imply that climate is just as sensitive to percentage changes in upper tropospheric water vapor, where the mixing ratio is very small, as it is to percentage changes in the planetary boundary layer, which contains the bulk of total column water vapor. Hence, increases in total column water vapor with temperature are not necessarily indicative of a positive water vapor feedback in climate; the changes in the upper troposphere, very small in terms of mixing ratio, are critically important.

6. CONCLUSION

A simple representation of the climate system--in which key processes are assumed to have a linear effect on system response and to be independent of each other--is used to show the extent to which climate response to an external radiative perturbation depends upon feedback processes. Examination of the results of GCM simulations of CO_2 doubling shows wide differences in cloud feedback amongst models. The conclusion drawn is that clouds *may* have a strong influence on climate change, but we are far from knowing the magnitude, and perhaps even the sign, of this influence.

GCM models are consistent in having a large positive feedback associated with water vapor, in that atmospheric water vapor increases with increasing temperature, thereby amplifying the effect of any external forcing. However, upper tropospheric water vapor depends on moist convection and precipitation--processes that may not be well parameterized in the models used for these studies. A study with a one-dimensional radiative equilibrium model with convective adjustment shows that a percentage change in upper tropospheric water vapor, where the mixing ratio is very small, is just as important as a percentage change in the planetary boundary layer, which contains the bulk of total column water vapor. Hence, increases in total column water vapor with temperature are not necessarily indicative of a positive water vapor feedback; the changes in the upper troposphere, very small in terms of mixing ratio, are critically important. Therefore, if the Lindzen hypothesis is correct, that the enhanced convection associated with radiative warming will lead to a drier upper troposphere, then water vapor feedback could be much smaller than in present models, and may even be negative.

ACKNOWLEDGEMENT

The author would like to thank Dr. William Ridgway for helpful discussions on radiative transfer methods and for modifying the radiation codes, supplied by Dr. Ming-Dah Chou, so that they could be called as external routines by the radiative-convective model used in this investigation.

REFERENCES

Chou, M.-D., and A. Arking, 1980: Computation of infrared cooling rates in the water vapor bands, *J. Atmos. Sci.*, **37**, 855-867.

Chou, M.-D., and A. Arking, 1981: An efficient method for computing the absorption of solar radiation by water vapor, *J. Atmos. Sci.*, **38**, 798-807.

Chou, M.-D., and L. Peng, 1983: A parameterization of the absorption in the 15-μm CO_2 spectral region with application to climate sensitivity studies, *J. Atmos. Sci.*, **40**, 2183-2192.

Cess, R.D., G.L. Potter, J.P. Blanchet, G.J. Boer, S.J. Ghan, J.T. Kiehl, H. Le Treut, Z.-X. Li, X.-Z. Liang, J.F.B. Mitchell, J.-J. Morcrette, D.A. Randall, M.R. Riches, E. Roeckner, U. Schlese, A. Slingo, and K.E. Taylor, 1989: Interpretation of cloud-climate feedback as produced by 14 atmospheric general circulation models, *Science*, **245**, 513-516.

Hansen, J. A., Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner, 1984: Climate sensitivity: analysis of feedback mechanisms. *Climate Processes and Climate Sensitivity, Geophys. Monogr. Ser., 29*, Ed. by J. E. Hansen and T.Takahashi, 130-163, AGU, Washington, D.C.

Lindzen, R.S., 1990: Some coolness concerning global warming, *Bull. Amer. Meteor. Soc.*, **71**, 288-299.

Mitchell, J.F.B., C.A. Senior, and W.J. Ingram, 1989: CO_2 and climate: a missing feedback? *Nature*, **341**, 132-134.

Ramanathan, V., and J.A. Coakley, Jr., 1978: Climate modeling through radiative-convective models, *Rev. Geophys. and Space Phys.*, **16**, 465-489.

Wetherald, R.T., and S. Manabe, 1988: Cloud feedback processes in a general circulation model, *J. Atmos. Sci.*, **45**, 1397-1415.